Feedbacks between the isentropic slope and wave generation at tropospheric levels

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Abstract

The authors examine feedbacks between the isentropic slope and the generation of eddy activity in the Northern Hemisphere winter. The results reveal that periods of enhanced lower tropospheric eddy heat fluxes (and thus wave generation) are preceded by anomalously steep isentropic surfaces in the free troposphere, and followed by anomalously poleward eddy momentum fluxes near the tropopause. The lag between the isentropic slope and wave generation is consistent with our theoretical understanding of baroclinic instability. The lag between the eddy fluxes of heat and momentum suggests that variability in the momentum fluxes is driven in part by variability in wave generation near the Earth’s surface.

The observed feedbacks between the isentropic slope and the generation of eddy activity in the lower troposphere have implications for the response of the extratropical circulation to a range of forcings. Here the authors examine the implications for the tropospheric response to stratospheric variability. It is demonstrated that the mass circulation anomalies associated with variations in the stratospheric flow account for a component of the observed changes in the tropospheric isentropic slope and lower tropospheric wave generation, but do not account for the observed changes in the wave fluxes of momentum at the tropopause level.
Context

Feedbacks between the mean-flow and the wave fluxes of heat and momentum play a fundamental role in the general circulation of the extratropical troposphere (e.g., Holton 2004 Ch. 10; Vallis 2006 Ch. 12). For example, wave activity is predominantly generated near the surface in regions of large isentropic slopes (Figure 1a; e.g., Stone 1978). Much of the wave activity generated in the lower troposphere propagates vertically and is thus associated with poleward fluxes of heat in the free troposphere. If the wave activity dissipates in the upper troposphere at roughly the same latitude range as the wave source (Figure 1b), then the westward torque aloft drives a residual (mass) circulation with rising motion equatorward of the wave source and sinking motion poleward of the wave source. The anomalous vertical motion acts to reduce the lower tropospheric isentropic slope that generated the wave activity in the first place. As such, the residual circulation induced by the wave breaking aloft acts as a negative feedback that attenuates lower tropospheric baroclinicity.

A somewhat different relationship holds between the lower tropospheric isentropic slope and the wave fluxes of momentum. For example, assume that a component of the upper tropospheric wave activity in Figure 1b propagates equatorward near the tropopause. The equatorward wave propagation is associated with poleward momentum fluxes that converge at the source latitude (Figure 1c). Viewed in isolation, the momentum fluxes induce a residual circulation that acts to reinforce rather than damp the lower tropospheric isentropic slope in the wave source region (Figure 1c). Hence, the residual circulation induced by the momentum flux convergence aloft acts as a positive feedback that increases lower tropospheric baroclinicity.
In the climatological-mean, the wave driving due to the vertical convergence of the wave flux (i.e., the vertical convergence of the heat flux) is larger than that due to the meridional divergence of the wave flux (i.e., the meridional convergence of the momentum flux) (e.g., Edmon et al. 1980). Hence, the total wave driving is westward at the tropopause level, but the momentum fluxes spread the wave driving over a wider range of latitudes than the wave source (Figure 1d). Since the momentum fluxes spread the wave driving meridionally, it follows that they also reduce the amplitude of the compensating poleward mass flux at the latitude of the wave source. As a result, the wave-driven residual circulation attenuates the lower tropospheric isentropic slope, but not as much as it would if all the wave-breaking was concentrated at the latitude of the wave source (see also Robinson 2000, 2006).

The feedbacks between the isentropic slope, tropospheric wave generation and tropospheric wave dissipation reviewed schematically in Figure 1 are known to play a key role in determining the long-term mean extratropical circulation. For example, the largest eddy fluxes of heat are roughly collocated with the regions of largest lower tropospheric baroclinicity (Kushner and Held 1998). However, the role of the isentropic slope in driving variability in the extratropical wave fluxes of heat and momentum remains unclear. The amplitude of such a feedback is important, since robust forcing of tropospheric eddies by variations in the isentropic slope is theorized to play a potentially key role in driving: 1) the internal feedbacks that give rise to the annular modes (e.g., Robinson 2000, 2006; Lorenz and Hartmann 2001, 2003); 2) the storm track response to extratropical sea-surface temperature anomalies (e.g., Kushnir et al. 2002 and references therein; Brawshaw et al. 2008); 3) the extratropical tropospheric response to
stratospheric variability (e.g., Song and Robinson 2004); and 4) the storm track
response to increasing greenhouse gases (e.g., Kushner et al. 2001; Yin 2005; Frierson
2006; Lu et al. 2008, 2010; Chen et al. 2010; O’Gorman 2010; Scaife et al. 2011; Butler
et al. 2011).

The primary purpose of this study is to examine the observational evidence for
feedbacks between the isentropic slope and the eddy fluxes of heat and momentum in
the Northern Hemisphere tropospheric circulation. The secondary purpose is to
examine the relevance of such feedbacks for the dynamics of stratosphere/troposphere
coupling. The results are presented in two sections. In the first results section, we
provide observational evidence that suggests variability in the tropospheric isentropic
slope drives robust changes in wave generation in the lower troposphere. In the second
results section, we review the implications of the linkages between the isentropic slope
and wave generation for a range of problems in climate dynamics, and argue that the
linkages explain at least a component of the tropospheric response to stratospheric
variability. The dataset is discussed in Section 2. Results are presented in Sections 3 and
4. Conclusions are given in Section 5.

2. Data and analysis

The primary dataset is 4x daily values of the ERA-Interim reanalysis from
January 1, 1989 to December 31, 2009 (ECMWF Newsletter 110; Dee et al. 2011). The
reanalyses products are available on a 1.5 degree x 1.5 degree mesh, and are daily and
zonally averaged before computing regressions and correlations. The zonal-mean fluxes
of heat and momentum are defined as $[\mathbf{v}^* T^*]$ and $[\mathbf{u}^* \mathbf{v}^*]$, respectively, where brackets
denote the zonal-mean and stars denote the departure from the zonal-mean. The fluxes are calculated at $4x$ daily resolution before being daily averaged. Anomalies are defined throughout the study as departures from the long-term mean seasonal cycle. All results are calculated for the Northern Hemisphere and for the winter months of December, January, February (DJF).

We exploit the ERA-Interim reanalysis products in both isentropic and pressure coordinates. To orient the reader to our use of both coordinate systems, Figure 2 shows the DJF, zonal-mean Northern Hemisphere circulation as a function of latitude/pressure (left) and latitude/potential temperature (right). In both panels, the height of the tropopause (bold line) is approximated as the level where potential vorticity is equal to 2 potential vorticity units.

We will refer to Figure 2 throughout the study. Key surfaces used in the results section include: 1) The 300 K isentrope, which extends to the surface in the tropics, spans the free troposphere at middle latitudes, and reaches the tropopause at polar latitudes; 2) The 700 hPa level, which lies near the top of the atmospheric boundary layer; and 3) The 200 hPa level, which lies in the upper troposphere/lower stratosphere region.

Variability in the strength of the extratropical stratospheric vortex is defined as the leading principal component time series of the zonal-mean 10 hPa geopotential height field. The index is generated according to Baldwin and Thompson (2009), and is referred to as the Northern annular mode index at 10 hPa (NAM$_{10}$). By definition, positive values of the NAM$_{10}$ index denote a stronger than normal stratospheric vortex, and vice versa.
Statistical significance of key results is assessed using a one-tailed test of the t-statistic. The effective degrees of freedom are estimated using the autocorrelation characteristics of the respective time series. By definition, standardized time series have a mean of zero and standard deviation of one.

3. Feedbacks between the isentropic slope and eddy fluxes at tropospheric levels

The meridional slope of extratropical isentropic surfaces provides a succinct measure of the tendency of the atmosphere to generate wave activity (e.g., Stone 1978; Stone and Nemet 1996). The isentropic slope is - by definition - given as (minus) the ratio of the meridional and vertical gradients in potential temperature. Increases in the meridional potential temperature gradient (i.e., the baroclinicity) and decreases in the vertical potential temperature gradient (i.e., the static stability) both increase the isentropic slope and thus the instability of the flow to the generation of baroclinic waves. The meridional slope of the isentropes is related to the available potential energy of the general circulation (e.g., Lorenz 1955) and is closely related to the Eady growth rate (the Eady growth rate is proportional to the meridional temperature gradient divided by the buoyancy frequency; Lindzen and Farrell 1980).

The isentropic slope is calculated here as the meridional gradient in pressure along isentropic surfaces, that is:

\[ \frac{\frac{\partial \theta}{\partial y}}{\frac{\partial \theta}{\partial p}} \frac{\partial p}{\partial y} \equiv s_p \]
where $\theta$ denotes potential temperature, $p$ denotes pressure, and the subscripts denote the variable that is held constant in the derivative. The resulting slope ($s_p$) is in units of hPa/km and is converted to m (vertical)/km (horizontal) using the hypsometric equation assuming a scale height of 7 km and a reference pressure of 1000 hPa.

The isentropic slope affects wavelike variability not only through its relation to baroclinic instability, but also through its projection onto the meridional gradient of potential vorticity (PV) along isentropic surfaces. For example, if we neglect the relative vorticity contribution to zonal-mean PV, the meridional gradient in PV along isentropic surfaces can be written as:

$$\left( \frac{\partial P}{\partial y} \right)_\theta = P \left( \frac{\beta}{f} - \frac{s_p}{\partial p} \right)$$

where $P$ denotes the zonal-mean potential vorticity. Hence the vertical structure of the isentropic slope is directly related to the meridional PV gradient which, in turn, determines the index of refraction for wave propagation.

Figures 3-5 examine the linkages between variability in the isentropic slope and the eddy fluxes of heat at 700 hPa averaged over the mid-high latitudes of the Northern Hemisphere (50-90°N). The eddy fluxes of heat at 700 hPa are a proxy for the eddy fluxes of PV at the lower boundary (i.e., the divergence of the eddy heat flux at the surface). Thus variations in the resulting time series (hereafter $[v^*T^*]_{50-90\text{°N}, 700\text{hPa}}$)
approximate variations in the generation of wave activity in the lower troposphere poleward of 50°N.

Figure 3 shows (top) the isentropic slope averaged between 285 K and 300 K, (middle) zonal-mean temperatures at 700 hPa, and (bottom) the eddy fluxes of momentum at 200 hPa regressed onto standardized values of $[v^*T^*]_{50-90^\circ N, 700 hPa}$ as a function of lag and latitude. The shading in all panels indicates the zonal-mean eddy heat fluxes at 700 hPa regressed onto $[v^*T^*]_{50-90^\circ N, 700 hPa}$. In all cases, lags less than zero denote results that precede peak amplitude in $[v^*T^*]_{50-90^\circ N, 700 hPa}$, and vice versa. Note that by construction, the heat flux regressions peak at lag 0 and at middle-high latitudes.

The results in Figure 3a reveal a distinct pattern of tropospheric isentropic slope anomalies both before and after peak amplitude in mid-high latitude wave generation. The ~5-10 day period before peak wave generation is marked by robust increases in the isentropic slope poleward of 50°N; the period following peak wave generation is marked by similarly robust decreases in the isentropic slope over middle and high latitudes. Hence as summarized schematically in Figures 1a and 1b, periods of anomalous tropospheric wave generation are a) preceded by anomalously steep isentropic slopes consistent with anomalous instability of the mean flow and b) followed by anomalously flat isentropic surfaces consistent with the stabilizing effect of the resulting residual circulation. Note that both the eddy heat fluxes and the response of the isentropic slope propagate poleward with time, consistent with the observed poleward propagation of anomalies in the extratropical zonal-mean flow (Feldstein 1998).
The changes in the isentropic slope in Figure 3a primarily reflect changes in the
temperature field at polar latitudes. Figure 4 shows zonal-mean pressure on isentropic
surfaces regressed onto standardized values of the $[v^*T^*]_{50-90^\circ N, 700 hPa}$ index time series as a
function of isentropic level and latitude, averaged over lags -10 to 0 days (top panel) and
0 to +10 days (bottom panel). The 10 day period before peak wave generation is marked
by anomalously low pressure on isentropic surfaces at polar latitudes (Figure 4 top)
while the 10 day period after peak wave generation is marked by pressure anomalies in
the opposite sense (Figure 4 bottom). As expected, the changes in polar pressure in
isentropic coordinates are accompanied by similarly signed changes in polar
temperatures at the 700 hPa level (Figure 3b). The pronounced out-of-phase
relationship in polar temperature anomalies (and thus in mid-high latitude isentropic
slope anomalies) between the periods before and after peak amplitude in the heat fluxes
is not a statistical artifact of the (quasi-random) periodicity inherent in the eddy forcing
(Appendix A).

The regression of the eddy momentum flux onto standardized values of
$[v^*T^*]_{50-90^\circ N, 700 hPa}$ (Figure 3c) reveals that periods of anomalous wave generation in the
mid-high latitude lower troposphere are followed by anomalous poleward momentum
fluxes (and thus anomalous equatorward wave activity fluxes) at the tropopause level
(Figure 3c). The momentum fluxes lag and are shifted equatorward of the region of
largest lower tropospheric heat fluxes. Hence the results are broadly consistent with the
“lifecycle paradigm” of tropospheric waves (e.g., Simmons and Hoskins 1978): i.e.,
waves are generated in the lower troposphere and propagate vertically over a period of
several days to the tropopause level, where a component of the wave activity propagates equatorward (Simmons and Hoskins 1978; see also Figure 1). The anomalous momentum fluxes in panel c) influence the isentropic slope through the attendant changes in the residual circulation (e.g., Figure 1c), and this influence is evidenced as the relatively rapid damping of the slope anomalies in Figure 3a at positive lags. The influence of the momentum fluxes on lower tropospheric baroclinicity is readily apparent in regressions analogous to those shown in Figure 3 but based on an index of the wave fluxes of momentum averaged over the mid-high latitudes (not shown).

The key results in Figures 3 are summarized in Figure 5. Periods of anomalous wave generation are 1) preceded by anomalously steep isentropic surfaces in the mid-high latitude lower troposphere and 2) followed by an increase in the momentum fluxes at 200 hPa and anomalously flat isentropic surfaces. The relationships highlighted in Figure 5 are significant at the 95% level (Figure 5 caption). They are readily reproducible in regressions based on indices of the isentropic slope and/or wave fluxes of momentum averaged over mid-high latitudes (not shown). And as noted above they are not due to the (quasi-random) periodicity inherent in the eddy forcing (Appendix A).

The results in this section are consistent with basic large-scale dynamics. The precursor in the isentropic slope is consistent with the dynamics of developing baroclinic waves (Stone 1978; Stone and Nemet 1996) and with the close relationship between available potential energy and the amplitude of the wave component of the flow (e.g., O’Gorman 2010). A causal linkage between the isentropic slope and the eddy fluxes of heat has been demonstrated in the context of the long-term mean circulation (e.g., Kushner and Held 1998) and is presumed to underlie the dynamics of the annular
modes (e.g., Robinson 1991, 2000). But to our knowledge a causal linkage between variations in the isentropic slope and the eddy fluxes of heat has not been demonstrated in the context of observed atmospheric variability. The flattening of the isentropic slope after peak amplitude in the heat fluxes is consistent with the effects of the residual circulation driven by the wave-breaking and meridional propagation aloft. The several day lag between the eddy fluxes of heat and momentum is expected from the “lifecycle” of vertically and meridionally propagating tropospheric waves.

In the next section we review the implications of the observed linkages between the isentropic slope and eddy generation for tropospheric climate variability, and explore the potential role of the linkages in stratosphere/troposphere coupling.

4. General implications and role in stratosphere/troposphere coupling

There are several obvious implications for a robust linkage between variability in the isentropic slope and extratropical tropospheric wave generation. For example:

1) The annular modes are theorized to derive in part from feedbacks between the momentum fluxes aloft, lower tropospheric baroclinicity, and the generation of baroclinic waves (e.g., Robinson 2000; Lorenz and Hartmann 2001, 2003). The annular mode dynamics envisioned in Robinson (1991) and the corresponding feedback loop articulated in Robinson (2000) hinge on the forcing of anomalous wave activity by changes in the lower tropospheric isentropic slope. The results in the previous section provide observational support for such forcing.

2) A range of forcings are predicted to drive changes in the storm tracks via their influence on the baroclinicity and thus isentropic slope in the extratropics. Increasing
greenhouse gases are predicted to steepen the extratropical isentropic slope via changes in both the horizontal temperature gradient and static stability (e.g., Kushner et al. 2001; Yin 2005; Frierson 2006; Lu et al. 2008, 2010; Chen et al. 2010; O’Gorman 2010; Scaife et al. 2011; Butler et al. 2011); Antarctic ozone depletion is predicted to steepen the isentropic slope at southern high latitudes due to the radiative effects of the Antarctic ozone hole (Grise et al. 2009); and midlatitude sea-surface temperatures anomalies are predicted to perturb surface baroclinicity in the vicinity of the extratropical storm tracks (e.g., Kushnir et al. 2002 and references therein; Brayshaw et al. 2008). The results in the previous section provide observational support for a robust response in tropospheric wave generation to such changes in the isentropic slope.

3) Wave-driven variability in the stratospheric circulation is expected to drive changes in the tropospheric isentropic slope due to the downward penetrating mass circulation (e.g., Haynes et al. 1991; Song and Robinson 2004; Thompson et al. 2006). The influence of stratospheric processes on the tropospheric isentropic slope is examined in more detail in this section.

Wave-driven variability in the strength of the extratropical stratospheric vortex is linked to similarly signed changes in the tropospheric circulation that persist for at least a month (Baldwin and Dunkerton 2001). The extratropical tropospheric response maps onto the structure of the tropospheric “annular modes” of climate variability (Thompson and Wallace 2000), and thus onto tropospheric wave fluxes of heat and momentum reviewed in the previous section. In the case of sudden stratospheric warmings, the changes in the tropospheric flow project onto the “low” index polarity of the annular mode. As such, the tropospheric response is marked by: 1) Weaker than normal
eastward flow near 55-60°N juxtaposed against stronger than normal eastward flow near 35-40°N; 2) Weaker than normal poleward eddy momentum fluxes at the tropopause level near 45-50°N; and 3) Weaker than normal poleward eddy heat fluxes (and hence weaker than normal wave generation) near the surface near 55-60°N (Limpasuvan et al. 2004).

The prevailing mechanisms thought to underlie stratosphere/troposphere coupling include: 1) The so-called balanced or “downward control” response (Haynes and Shepherd 1989; Haynes et al. 1991; Thompson et al. 2006), i.e., geostrophic and hydrostatic adjustment to anomalous wave driving at stratospheric levels; 2) Feedbacks between the stratospheric circulation and the direction and/or speed of wave propagation in the upper troposphere/lower stratosphere (e.g., Chen and Robinson 1992; Perlwitz and Harnik 2003; Kushner and Polvani 2004; Wittman et al. 2007; Chen and Held 2007; Chen et al. 2008; Simpson et al. 2009; Shaw et al. 2010); and 3) Feedbacks between the downward penetrating stratospheric circulation and the generation of wave activity in the lower troposphere (e.g., Song and Robinson 2004).

All the above mechanisms may contribute to observations of stratosphere/troposphere coupling. The results in Figures 6-8 suggest that the feedbacks outlined in Section 3 also explain at least a component of the observed coupling. Figure 6 shows latitude/isentrope plots of select zonal-mean variables regressed onto inverted values of the NAM$_{10}$ index (the index is multiplied by -1 so that the regressions correspond to conditions consistent with sudden stratospheric warmings). The regressions are calculated for daily-mean data, and are averaged over lags extending from ten days before to ten days after peak amplitude in the NAM$_{10}$ index. Figure 7 shows analogous
regressions, but for results at the 300 K surface as a function of lag and latitude. Recall that the 300 K surface spans the middle and upper troposphere at mid-latitudes (Figure 2).

The top panels in Figures 6 and 7 show results for the zonal-mean zonal wind. As noted in numerous previous studies (e.g., Thompson and Wallace 2000; Baldwin and Dunkerton 2001; Limpasuvan et al. 2004), westward anomalies in the polar stratospheric flow (weakenings of the stratospheric vortex) are coupled with westward anomalies centered ~55°N and eastward anomalies centered ~35°N degrees that extend to the surface of the Earth (Figure 6a). The tropospheric wind anomalies peak during the ~2-3 week period following peak amplitude in the stratospheric flow (Baldwin and Dunkerton 2001; Figure 7a).

The middle panels in Figures 6 and 7 show results for the zonal-mean pressure on isentropic surface. Variability in zonal-mean pressure on isentropic surfaces provides a simple measure of variability in the zonal-mean diabatic circulation (i.e., the residual mass circulation): e.g., a rise in pressure on a given isentropic surface is equivalent to a rise in potential temperature on the nearby pressure surfaces. During weakenings of the stratospheric flow, the anomalous residual circulation advects isentropic surfaces towards higher pressure (downwards) over the polar cap. As such, the anomalous residual circulation should be reflected as an increase in pressure on polar isentropic surfaces.

The results in Figure 6b confirm that weakenings of the stratospheric vortex are accompanied by rising pressure on isentropic surfaces that extend throughout the depth of the polar atmosphere. The pressure anomalies are nearly constant with height down
to ~285 K (~700 hPa) at 65°N, which suggests that the downward mass flux is also roughly constant with height from the middle stratosphere to the lower troposphere at these latitudes. The polar tropospheric pressure anomalies peak during the ~2-3 week period following largest amplitude in the stratospheric flow (Figure 7b). The damping of the pressure signal in the high latitude polar troposphere and the weak pressure anomalies at 25°N and 50°N (Figure 6b) are consistent with the competing effects of the tropospheric wave fluxes of heat and momentum on the residual circulation (see Appendix B).

The bottom panels in Figures 6 and 7 show the corresponding results for the isentropic slope. The rises in polar pressure evident in Figures 6b and 7b indicate a flattening of the isentropic surfaces at middle and subpolar latitudes, as evidenced in Figures 6c and 7c. Thus the downward motion over the polar cap associated with weakenings of the stratospheric flow leads to a reduction in the isentropic slope not only at stratospheric levels but at tropospheric levels as well (Figures 6c and 7c). The reductions in tropospheric isentropic slope span most of the extratropics and exhibit largest amplitude following lag zero (Figure 7c). The results in the middle and bottom panels of Figures 6 and 7 are consistent with geostrophic and hydrostatic balance with the wind anomalies in the top panels. But they nevertheless make explicit the role of the residual circulation in perturbing lower tropospheric baroclinicity.

Are the changes in the tropospheric isentropic slope evidenced in Figure 7c associated with changes in wave generation, as inferred from the results in Section 3? Figure 8a shows the anomalous eddy heat fluxes at 700 hPa (contours) and the isentropic slope at 300 K (shading) regressed onto inverted values of the NAM10 index.
(The isentropic slope results are reproduced from Figure 7c). Consistent with the composites shown in Limpasuvan et al. (2004), weakenings of the polar stratospheric vortex are associated with a reduction in the eddy fluxes of heat in the mid-high latitude lower troposphere. The decreases in the eddy fluxes of heat occur over roughly the same latitude range as (and generally lag) the largest decreases in the tropospheric isentropic slope (Figure 8a). The anomalous tropospheric heat fluxes are thus downgradient (i.e., they are decreased in regions of decreased baroclinicity) and consistent with forcing of the eddy fluxes of heat by the anomalous tropospheric isentropic slope (Section 3).

The changes in the tropospheric eddy fluxes of momentum are more difficult to interpret. The shading in Figure 8b shows the regression of the eddy fluxes of momentum at 200 hPa onto inverted values of the NAM\textsuperscript{10} index. As documented in Limpasuvan et al. (2004), weakenings of the stratospheric vortex are associated with anomalous equatorward momentum fluxes that peak around 55°N a few days before peak amplitude in the NAM\textsuperscript{10} index. A partial explanation for the changes in the momentum fluxes lies in the “lifecycle paradigm” of tropospheric baroclinic waves evidenced in Figure 3c. In that figure, a ~5 K m/s change in the heat fluxes at 700 hPa centered near 60°N is associated with a ~5 m\textsuperscript{2}/s\textsuperscript{2} change in the momentum fluxes at 200 hPa centered near 50°N. The heat flux anomalies in Figures 8a and 8b peak near 1 K m/s and are only notable at positive lags. Thus the linear relationship between the wave fluxes of heat and momentum shown in Figure 3c explains only about a quarter of the changes in the momentum fluxes after day +5 and none of the large changes in the momentum fluxes at negative lag.
The large changes in the momentum fluxes that occur several days before peak amplitude in the stratospheric flow may reflect feedbacks between the zonal-flow and wave propagation at the tropopause level, which have not been considered here (e.g., Chen and Robinson 1992; Kushner and Polvani 2004; Wittman et al. 2007; Chen and Held 2007; Chen et al. 2008). However, the largest changes in the momentum fluxes precede by several days the largest changes in the flow at 10 hPa. For this reason, we believe that it is equally plausible that the pulse of momentum fluxes at negative lags reflects the extension of the anomalous poleward focusing of wave activity to the tropopause level. Such focusing is a key aspect of a developing sudden warming at stratospheric levels. That the largest anomalous equatorward fluxes of momentum (poleward fluxes of wave activity) are concurrent at both tropospheric and lower-mid stratospheric levels (e.g., Thompson et al. 2006, c.f. Fig. 9) suggests that such focusing occurs simultaneously across a range of upper tropospheric and stratospheric levels. As such, the momentum fluxes at negative lags may be viewed as an integral component of a developing sudden warming and not necessarily the tropospheric response to it. Support for this hypothesis is beyond the scope of the current study.

5. Discussion

The influence of the mean flow on the eddy fluxes of heat and momentum can be viewed in the context of two processes: 1) the influence of the mean flow on the direction and dissipation of wave propagation in the free troposphere and 2) the influence of the mean flow on the generation of wave activity near the Earth’s surface. Both are clearly important in Earth’s atmosphere. The configuration of the mean flow...
determines the “index of refraction” for wave propagation and the “critical” regions for wave breaking. The tropospheric isentropic slope affects the production of wave activity in the lower troposphere as well as the characteristics of wave propagation.

Here we have focused on the slope of the isentropic surfaces as a measure of the baroclinicity and the static stability of the flow. The isentropic slope is related to the potential energy available to generate atmospheric eddies (e.g., Lorenz 1955; O’Gorman 2010), is of fundamental importance for baroclinic wave generation (e.g., Stone 1978; Stone and Nemet 1996), is intimately related to the growth rate in theories of baroclinic instability (e.g., Lindzen and Farrell 1980), and is directly related to the meridional PV gradients and thus the index of refraction for wave propagation (Section 3). The results in Section 3 confirm two expectations based on theoretical expectations: 1) periods of increased isentropic slope (and thus increased available potential energy) lead to enhanced wave generation near the surface and thus to enhanced wave dissipation at upper levels; and 2) the enhanced wave dissipation at upper levels drives a stronger residual circulation that acts to reduce the isentropic slope, providing a negative feedback between the mean flow (in this case the isentropic slope) and the waves. A causal linkage between the baroclinicity and the eddy fluxes of heat is readily apparent in the long-term mean (e.g., Kushner and Held 1998). The observations presented here provide observational evidence that such a linkage is also apparent in the context of atmospheric variability.

The observations in Section 3 also reveal that variability in wave generation in the lower troposphere precedes changes in the eddy fluxes of momentum at the tropopause level by several days, and that the momentum fluxes, in turn, force changes in lower
troospheric baroclinicity. The lag between the eddy fluxes of heat in the lower
troposphere and of momentum at the tropopause level is expected based on the lifecycle
view of baroclinic instability (e.g., Simmons and Hoskins 1978), and it suggests that a
component of the variability in the momentum fluxes is driven by changes in wave
generation in the lower troposphere.

The seemingly robust linkage between anomalies in tropospheric baroclinicity
and wave generation has implications for a range of climate processes. For example, the
dynamics thought to underlie the annular modes hinge in part on the forcing of eddy
generation by changes in the isentropic slope (e.g., Robinson 1991, 2000; Lorenz and
Hartmann 2001, 2003). The results shown here suggest that such forcing is a robust
aspect of observed lower tropospheric wave/mean flow interaction. Previous studies
have noted changes in baroclinicity (either via changes in the horizontal temperature
gradient or static stability) in simulations of anthropogenic climate change (e.g.,
Kushner et al. 2001; Yin 2005; Frierson 2006; Lu et al. 2008, 2010; Chen et al. 2010;
O’Gorman 2010; Scaife et al. 2011; Butler et al. 2011), sea-surface temperature
anomalies (e.g., Kushner et al. 2002 and references therein; Brayshaw et al. 2008), and
Antarctic ozone depletion (Grise et al. 2009). The results shown here provide
observational support for a causal linkage between such changes in baroclinicity and the
eddy fluxes of heat that characterize the extratropical storm tracks.

We have focused on zonal-mean relationships in this study. In practice, the
feedbacks between the isentropic slope and eddy fluxes of heat highlighted here derive
primarily from the sector of the hemisphere stretching from eastern North America to
the central North Atlantic (not shown). As such, the feedbacks are most pronounced
over the North Atlantic storm track region and as inferred earlier, are consistent with
the response of synoptic-scale waves to changes in baroclinicity. We have also focused
on the eddy fluxes of heat averaged poleward of 50°N. Similar results to those shown in
Figure 3 are derived using an index of the eddy fluxes of heat averaged poleward of
30°N. Interestingly, the precursor in the isentropic slope does not extend to the
anomalous heat fluxes equatorward of 50°N. The feedback between the isentropic slope
and the eddy fluxes of heat appears to be most pronounced for latitudes poleward of
50°N and over the Atlantic sector of the hemisphere.

The results also provide a partial explanation for the tropospheric response to
stratospheric variability. As shown in Section 4, periods of anomalous wave driving in
the stratosphere are marked by anomalous downward mass fluxes over the polar cap
that extend not only throughout the lower stratosphere but well below the tropopause.
The anomalous mass fluxes are evidenced as increases in pressure on polar isentropic
surfaces, and hence by a reduction in the isentropic slope at middle and high latitudes.
At tropospheric levels, the flattening of the isentropic surfaces is coincident with a
decrease in near-surface wave generation, i.e., anomalously equatorward lower
tropospheric wave heat fluxes. Hence stratospheric variability appears to perturb lower
tropospheric wave generation in a manner similar to that suggested in the numerical
results of Song and Robinson (2004): anomalous stratospheric wave drag drives an
anomalous residual circulation, and the anomalous residual circulation perturbs
tropospheric baroclinicity and thus wave generation.

That said, it is worth emphasizing that the linkages examined in Section 3
provide only a partial explanation for the tropospheric response to stratospheric
variability. The changes in the tropospheric heat fluxes account for only a fraction of the changes in the momentum fluxes after peak amplitude in the stratospheric flow, and for none of the changes in the eddy fluxes of momentum before peak amplitude in the stratospheric flow. It remains to be determined whether the anomalous momentum fluxes that precede peak changes in the stratospheric flow are: 1) an integral component of sudden stratospheric warmings (e.g., increased focusing of wave activity towards the pole at both stratospheric and upper tropospheric levels); and/or 2) a direct feedback between the zonal-flow and the eddy fluxes of heat and momentum at the tropopause level (e.g., Chen and Robinson 1992; Kushner and Polvani 2004; Wittman et al. 2006; and Chen and Held 20076; Chen et al. 2008).

The results in this study suggest that external forcing is more likely to project in a predictable manner onto the isentropic slope and wave generation than it is onto the meridional propagation of wave activity aloft. The results do not question the importance of the momentum fluxes for driving internal tropospheric variability. Rather, they suggest that the feedbacks between the isentropic slope and wave generation provide a robust framework for interpreting the storm track response to a range of external forcings.
Appendix A: Robustness of the results to periodicity in the eddy forcing.

Lag correlations between two time series may exhibit out-of-phase correlations at different lags due simply to the (quasi-random) periodicity inherent in the time series. The robustness of the feedbacks between the temperature field and the eddy fluxes of heat to (quasi-random) periodicity in the eddy forcing was examined as follows.

First, consider the pattern of observed 700 hPa temperatures regressed onto variations in the eddy fluxes of heat at middle to high latitudes. As shown in Figure 3b, 700 hPa temperatures exhibit large out-of-phase anomalies about lag zero that peak over the polar regions: periods before large heat fluxes are marked by anomalously cold polar temperatures (Figure 3b) and thus anomalously steep isentropes at mid-high latitudes (Figure 3a); periods following large heat fluxes are marked by anomalies in the opposite sense. The observed linkages between 1) 700 hPa temperatures averaged over the polar cap and 2) the convergence of the eddy fluxes of heat (also averaged over the polar cap) are summarized as the solid line in Figure A1. The amplitudes of the negative regression coefficients at negative lags are roughly the same as the amplitudes of the positive regression coefficients at positive lags.

Are the negative temperature anomalies before lag zero due to the periodicity inherent in the eddy forcing? To test this, we estimated the anomalous polar temperatures at 700 hPa due solely to forcing by the observed anomalous convergence of the eddy heat flux. This was done using a linearized form of the thermodynamic energy equation:
\begin{equation}
\frac{dT_{\text{Model}}}{dt} = H - \alpha T_{\text{Model}}
\end{equation}

where $T_{\text{Model}}$ is the (simulated) anomalous 700 hPa temperature, $H = -\frac{\partial}{\partial y} [\nu' T']$ is the observed convergence of the anomalous eddy heat flux averaged over 65-90°N, and $\alpha$ is the damping timescale. The damping timescale was estimated from the e-folding timescale of the observed temperatures at 700 hPa averaged over 65-90°N, and was found to be six days. Note that the damping term parametrizes both adiabatic (e.g., vertical motion) and diabatic (e.g., radiative fluxes) processes. The model was integrated over all days 1989-2009.

The dashed line in Figure A1 shows results derived by regressing 1) simulated polar cap temperatures ($T_{\text{Model}}$ in Equ. A1) onto 2) the observed convergence of the eddy fluxes of heat averaged over the polar cap (i.e., the same time series used to drive Equ. A1 and as a basis for the regressions given by the solid line in Figure A1). By construction, the simulated temperatures exhibit positive regression coefficients at lags when the temperature field is lagging the eddy forcing. The close correspondence at positive lags between regressions based on the observed and simulated temperatures indicates that Equ. A1 does a reasonable job of simulating the response of polar temperatures to the observed anomalous convergence of the eddy heat flux. If the eddy forcing exhibited sufficient periodicity to drive spurious negative correlations at negative lags, such negative correlations should appear in association with both the observed and simulated temperature time series. The absence of negative correlations
(let alone a negative peak in the correlations) before lag zero indicates that the periodicity inherent in the observed eddy forcing time series is not sufficient to yield large spurious negative correlations at negative lags.

The above results were repeated using as a basis of the regression the eddy heat fluxes averaged over 50-90°N (as used in Figure 3). The results are qualitatively unchanged. The above test was also repeated for a range of different latitude bands and levels. Overall, we found that the ratio between the regression coefficients at positive and negative lags (which is roughly \(-1\) one for the observations in Figure A1) is always at least three times larger for the observed temperatures than it is for temperatures calculated as the damped thermodynamic response to the observed convergence of the eddy heat flux.
Appendix B. Interpretation of the pressure response at tropospheric levels in Figure 6

As noted in Section 4, the high latitude pressure anomalies in Figure 6b are nearly uniform with height throughout the stratosphere but fall off with decreasing altitude in the polar troposphere. The polar stratospheric pressure anomalies indicate nearly constant downward mass flux anomalies above the tropopause. But the decrease in pressure anomalies in the polar troposphere suggests that the anomalous downward mass flux is attenuated by anomalous equatorward flow near the level of the tropopause. There are at least two possible explanations for the attenuation of the downward mass flux in the polar troposphere.

1) The attenuation is due to the transient response to stratospheric wave drag. For example, the balanced response to stratospheric wave drag includes poleward flow across the axis of the forcing. At steady-state, the compensating equatorward flow is limited to the surface and thus the downward mass flux over the pole is constant with height. But on relatively short time-scales, the compensating equatorward flow can occur at all atmospheric levels below the level of the forcing, and thus the downward mass flux decreases with decreasing altitude (e.g., Haynes and Shepherd 1989; Haynes et al. 1991; Holton et al. 1995). We view this explanation as unlikely since the attenuation should be visible at both stratospheric and tropospheric levels.

2) The attenuation of the pressure anomalies in the polar troposphere is due to the competing effects of the tropospheric eddy fluxes of heat on the anomalous residual circulation. For example, the pressure response in Figure 6b can be decomposed into
a “stratospheric component”, which is defined to be uniform with height from the stratosphere to the surface; and 2) a “residual”, which is defined as the difference between the total changes in pressure and the stratospheric component. Figure B1 shows the results of such a calculation where the stratospheric component is defined as the pressure changes at 330 K. The stratospheric component can be interpreted as the changes in the mass field that would result if the residual circulation extended uniformly from 330 K to Earth’s surface. The residual can be interpreted as the signal coming from internal tropospheric dynamics.

The residual (Figure B1 bottom) is dominated by 1) pressure decreases that are largest at the polar surface and damp with height towards the tropopause and 2) a comparatively weak dipole in the pressure field centered ~30°N and 50°N at 300 K. Both features are consistent with the effects of internal tropospheric dynamics on the tropospheric residual circulation. As demonstrated in Section 3, increases in tropospheric wave generation lead to a reduction in the tropospheric isentropic slope (Figure 3a) and an increase in pressure on polar isentropic surfaces (Figure 4 bottom). Hence the decreases in tropospheric wave generation associated with weakenings of the stratospheric vortex (contours in Figure 8a) should lead to changes in polar tropospheric pressure that resemble (but are of opposite sign to) the results in the bottom panel of Figure 4. The similarity between the patterns in bottom panels in Figures 4 and B1 suggests that the tropospheric eddy heat fluxes are acting to attenuate the effects of the stratospheric mass circulation in the high latitude troposphere in a manner consistent with the hypothesis 2) outlined above.
The pattern of pressure rises and falls at 30°N and 50°N at the 300 K level (Figure B1 bottom) is consistent with the effects of the anomalous momentum fluxes on the residual mass circulation. The anomalous equatorward momentum fluxes in Figure 8b should induce a pattern of meridional circulation anomalies analogous to those shown in Figure 1c, but with the sign flipped (Figure 1c shows the response to anomalous poleward momentum fluxes). Thus the anomalous momentum fluxes in Figure 8b will induce rising motion (and thus pressure falls) ~50°N and sinking motion (and thus pressure increases) ~30°N in the free troposphere, as indicated in Figure B1 (bottom).
References


a) Regions of large isentropic slope are associated with enhanced wave generation

b) Wave breaking aloft acts to reduce the tropospheric isentropic slope

c) Meridional wave propagation away from the source latitude acts to reinforce the tropospheric isentropic slope

d) Combined effects of vertical and meridional wave propagation

Figure 1. Schematics of the interactions between the isentropic slope and the wave fluxes of heat and momentum. Schematics denote distinct dynamical situations and do not necessarily reflect a series of events. In all panels sloping lines indicate isentropic surfaces; curly lines indicate the direction of the wave fluxes; shading (stippling) denotes regions of EP flux convergence (divergence); and rounded rectangles indicate the residual circulation. See text for details.
Figure 2. Northern Hemisphere December, January, February (DJF) mean, zonal-mean circulation in pressure (left) and isentropic (right) coordinates. Dashed lines indicate isentropic surfaces (left) and pressure surfaces (right); light solid lines indicate the zonal wind; heavy solid line indicates the 2 PV units isoline and thus approximates the height of the dynamical tropopause. The isentropic levels on the right panel correspond to the isentropic levels available in the ERA Interim dataset.
Figure 3. Lag regressions on standardized values of the eddy fluxes of heat at 700 hPa averaged 50-90N. Results are shown as a function of lag and latitude. Shading in all panels denotes the eddy fluxes of heat at 700 hPa (in K m/s). Contours denote a) The isentropic slope averaged over 285K and 300K (contours at 20 m (vertical) / 1000 km (meridional)). b) Temperatures at 700 hPa (contours at 0.2 K). c) The eddy fluxes of momentum at 200 hPa (contours at 2.25 m^2/s^2). All results are based on zonal-mean, daily-mean anomaly data (i.e., the seasonal cycle has been removed from the data) for DJF. The zero contour is omitted in all panels.
Figure 4. Regressions on standardized values of the eddy fluxes of heat at 700 hPa averaged 50-90N. (top) Pressure anomalies averaged over lags -10 to 0 (preceding the heat flux index); (bottom) Pressure anomalies averaged over lags 0 to +10 (following the heat flux index). Pressure results are based on zonal-mean, daily-mean anomaly data for DJF. The solid line indicates the 2 PV units isoline and thus approximates the height of the dynamical tropopause.
Figure 5. (a) Regressions on standardized values of the eddy fluxes of heat at 700 hPa averaged 50-90N. (Pluses) The isentropic slope at 300K averaged 50-90N (ticks every 10 m/1000 km); (Solid) Eddy fluxes of heat at 700 hPa (ticks every 1 K m/s). (Dashed) Eddy fluxes of momentum at 200 hPa (ticks every 1 m^2/s^2). Values exceeding 1.7 tickmarks (pluses), 1.8 tickmarks (dashed) and 0.4 tickmarks (solid) are significant at the 95% level based on a one-tailed test of the t-statistic (the effective degrees of freedom are estimated based on the observed autocorrelation characteristics of the time series).
Figure 6. Regressions on inverted and standardized values of the NAM index at 10 hPa. The zonal-mean zonal wind (top); pressure on isentropic surfaces (middle); the isentropic slope (bottom). Results are based on zonal-mean, daily-mean anomaly data for DJF. Results are averaged over lags -10 days (preceding the NAM10 index) to +10 days (following the NAM10 index). The solid line indicates the 2 PV units isoline and thus approximates the height of the dynamical tropopause.
Figure 7. Regressions on inverted and standardized values of the NAM index at 10 hPa. The zonal-mean zonal wind at 300K (top); pressure at 300K (middle); the isentropic slope at 300K (bottom). Results are based on zonal-mean, daily-mean anomaly data for DJF. Negative lags denote results preceding the NAM10 index, and vice versa.
Figure 8. Regressions on inverted and standardized values of the NAM index at 10 hPa. (a) Shading: The isentropic slope at 300K. Contours: The eddy fluxes of heat at 700 hPa. Contours at 0.4 K m/s; the zero contour is omitted. (b) Contours: as in panel a). Shading: The eddy flux of momentum at 200 hPa. Results are based on zonal-mean, daily-mean anomaly data for DJF. Negative lags denote results preceding the NAM10 index, and vice versa.
Figure A1. Solid line: Observed temperatures at 700 hPa (averaged 65-90N) regressed onto standardized values of the observed convergence of the eddy heat flux at 700 hPa (averaged 65-90N). Dashed line: As in the solid line, but the 700 hPa temperatures at 700 hPa are calculated based on the damped thermodynamic response to the observed convergence of the eddy heat flux. See Appendix text for details.
Figure B1. As in the middle panel of Figure 6, but the results are decomposed into (top) the pressure anomalies at 330K (the stratospheric component of the pressure changes; the horizontal line denotes the 330K surface) and (bottom) the residual. Note that the sum of both panels is by construction identical to the results in the middle panel of Figure 6.